

Planetary boundary layer structure and air mass transport during the International Arctic Ocean Expedition 1991

By E. DOUGLAS NILSSON, *Department of Meteorology, Stockholm University, S-106 91, Stockholm, Sweden*

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ABSTRACT

During the International Arctic Ocean Expedition 1991 north of latitudes 70°N from August 1 to October 6 (late summer and early autumn), the dominant air mass origin was from the seas that surround the central Arctic Ocean, while land was a much less important source. This resulted in air temperatures near 0°C, fogs and extensive stratus layers. In the autumn, long residence time of the air over the pack ice and subsidence became more common, accompanied by decreasing cloudiness, foginess and air temperature. The residence time of air over the pack ice and the time since land contact had medians of 52 and 91 h, respectively. The structure and evolution of the marine boundary layer above the pack ice is described. A stable layer, 300–1700 m high, grew in height proportionally to the square root of the distance from the ice edge, the fetch, up to 3000 km. After 400 km, a mixed layer formed, typically 100–250 m thick, related to the presence of low level jets in 60–90% of the profiles. Layers of subcritical Richardson number were present in 65–80% of the profiles above and below the jet centres. Cloud streets and roll vortices were observed above the pack ice with wavelengths of 4–12 km. The high aspect ratios of 7–9 of the few earlier roll observations above ice were confirmed. Cold air advection and weak convection caused the roll vortices in aged air that had been advected usually 2000 km or more over the ice. Breaking of Kelvin–Helmholtz waves, gravity waves or other reasons for intermittent planetary boundary layer breakdown combined with vertical gradients in concentration are suggested to have caused the observed sudden changes, in an hour or less, in aerosol and tracer concentrations and to have influenced fog formation. Roll vortices are suggested as the cause of periodic changes in aerosol concentrations and fog formation.

1. Introduction

The International Arctic Ocean Expedition 1991 (IAOE-91) was a joint ship operation (Anderson et al., 1991) in the Barents sea, Greenland sea and central Arctic Ocean during August, September and the beginning of October 1991, see Fig. 1, in which the Swedish ice-breaker *Oden*, the German ice-breaking research vessel *Polarstern* and the American ice-breaker *Polar Star* participated. On board the ice-breaker *Oden* an atmospheric programme was conducted which was designed to study the natural sulfur cycle of the Arctic Marine Boundary Layer (MBL) (Leck et al., 1996).

Concentrations of the measured sulfur com-

pounds depend on source strengths, chemical reactions and aerosol physical processes, but also on meteorological processes. The air mass origin and transformation during transport, clouds and fogs, the planetary boundary layer (PBL) structure and mixing processes are of fundamental importance and have to be understood if their influence is to be distinguished from that of chemical and physical aerosol processes. Especially our understanding of Arctic fog and cloud formation and of the Arctic PBL structure and processes at different conditions is far from complete. This paper therefore describes the PBL structure related to the weather situation and air mass origin. The evolution of the PBL structure with increased fetch during advection over the pack ice is also evaluated. Fog and

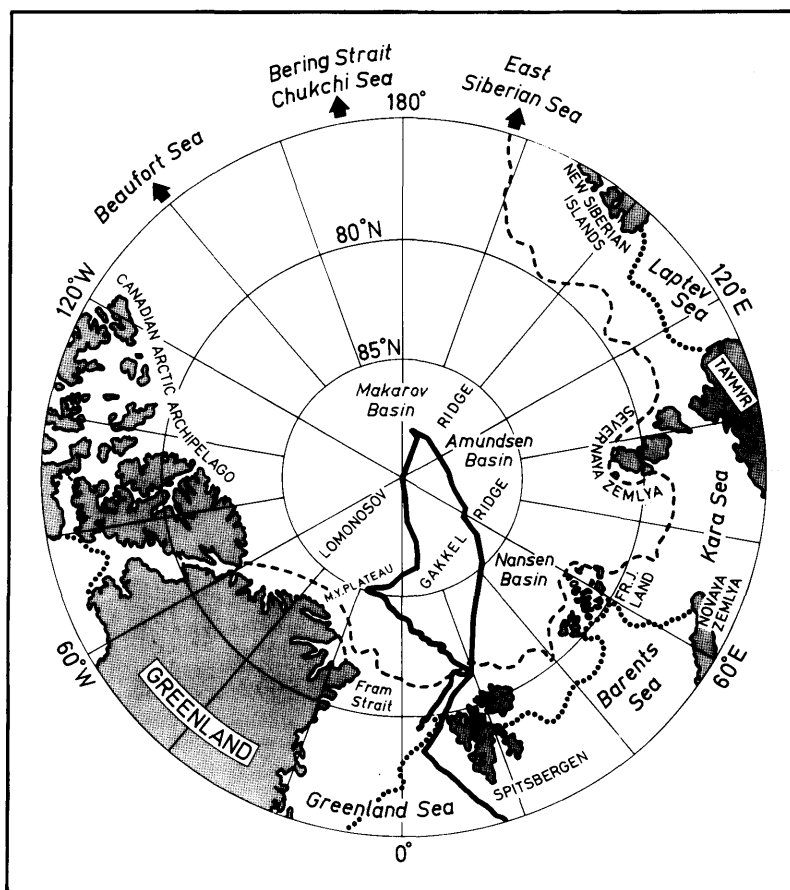


Fig. 1. Map of the Arctic Ocean with maxima and minima ice cover between 1 August and 6 October (Julian days 212–279) 1991 and the cruise track (followed counter clockwise) of the ice breaker *Oden*. The cruise began in Tromsø, northern Norway. *Oden* operated in open sea in the Barents Sea, the Greenland Sea, the Fram strait and north of Spitsbergen in the marginal ice zone until 7 August (Julian day 219), when it anchored in Isfjorden, Spitsbergen, for service (not included in the cruise track). From 17 August (Julian day 229), *Oden* operated in the pack ice covered central Arctic ocean. The north pole was crossed on 7 September (Julian day 250). On 5 October (Julian day 278), *Oden* reached open sea again in the Fram strait and the last atmospheric station took place in the Greenland Sea 6 October.

cloud observations are presented in an accompanying paper (Nilsson and Bigg, 1996).

2. Experimental approach

2.1. Surface observations

On board the icebreaker *Oden*, complete meteorological surface observations were made every hour (or every half hour during the stations of the atmospheric programme). Air temperature

(30 m asl), water temperature (3 m bsl), relative humidity (30 m asl), air pressure (reduced to surface level), relative wind speed (34 m asl), relative wind direction (34 m asl) were observed by a MARITEX station for off-shore use. The water temperature during stations was taken by a bucket thermometer. In addition visual observations of cloud type, cloud cover, fog, precipitation type, visibility and general weather were performed. The position, heading, speed and manoeuvres of the ship were also logged. During the stations of

the atmospheric programme, on occasions of rawinsonde ascents and for the 6 hourly SHIP observations, relative humidity was measured with an Assman psychrometer. At other times, relative humidity from the MARITEX station was used, corrected to the Assman psychrometer. We have used the temperature at 5 cm depth, measured with a Fluke thermometer immersion probe in ice cores by the remote sensing programme (Carlström and Ulander, 1993), as the ice surface temperature.

2.2. Rawinsoundings

Throughout the expedition, rawinsondes (Vaisala RS80-15N) were launched at 6–8 h intervals during the stations of the atmospheric programme to receive and record (Vaisala DigiCORA MW 11 Rawinsonde set and MF 12 microdisk recorder) the vertical distribution of the following parameters: pressure p ($\sigma_p = 0.5$ hPa corresponding to $\sigma_z \approx 4$ m for the height z), relative humidity RH ($\sigma_{RH} = 3\%$), air temperature T ($\sigma_T = 0.2^\circ\text{C}$), wind direction and wind speed U ($\sigma_U = 0.5$ m s⁻¹ in vector length). σ is the standard deviation of repeated calibrations, except for the wind for which it is the standard deviation of differences in a group of several ascents with two sondes suspended from one balloon and tracked by independent receiving systems. From the observed temperature, pressure and relative humidity we calculated the virtual temperature T_v and virtual potential temperature θ_v . The wind was measured with the Vaisala multifrequency composite VLF method for the Navaid wind-finding system. The system could receive all 8 Omega stations and all 3 Sigma stations most of the time. The DigiCORA software allows the user control over data smoothing and editing and includes a research mode which omits data filtering. Each rawinsonde was calibrated at 0% relative humidity and indoor temperature (Vaisala GC 22 ground check set) and to outdoor conditions, including wind speed and wind direction, the moment before launch. To increase the vertical resolution the ascent rate was limited to an average of 3 m s⁻¹, which with an instrumental record interval of 2 s for T , p , RH and 10 s for U corresponded to average vertical resolutions of 6 and 30 m respectively. The launch site was the helicopter deck of *Oden*, 15 m asl.

2.3. Trajectories

5 days receptor oriented trajectories have been calculated for the arrival level of 1000 hPa and arrival times at 00, 06, 12 and 18 UTC during the entire expedition, with the 1991 version of the 3-dimensional trajectory model of McGrath (1989) using the analysed horizontal and vertical wind components and mass fields of the ECMWF (European Center for Medium range Weather Forecasts). The analysed fields had a vertical distribution of 19 levels until 17 September 1991, after which ECMWF changed to 31 vertical levels. To avoid curvature problems in the vicinity of the pole the model switches to a rotated coordinate system for latitudes $>60^\circ$. The accuracy of the trajectory model is examined in the Appendix.

2.4. Satellite images and ice maps

From polar orbiting NOAA satellites images were received with a MacSatII Automatic Picture Transmission (APT) station. To follow changes in the density and extension of the Arctic pack ice, ice boundary maps prepared by NESDIS/Synoptic Analysis Branch (based on NOAA-11 and GOES-7 images) and ice maps prepared by the Norwegian meteorological institute (based on both satellite and ship observations) have been used.

3. Results and discussion

3.1. Weather during the IAOE-91

The weather during August was typical of the Polar Ocean and central Arctic Ocean summer when advection of air warmer than the melting point of sea ice contributes to the ice melting (Vowinkel and Orvig, 1970). Marine air dominated over both continental and subsided air masses. This gave ice and air temperatures close to 0°C , as seen in Fig. 2, and relative humidities near saturation since the advected air was cooled by the melting pack ice. The cooling led to a stably stratified PBL and formation of a dense and low deck of stratus clouds as well as advection fogs (Nilsson and Bigg, 1996). In late August, a family of cyclones entered from the Atlantic Ocean, reinforced a semi-permanent cyclone at the pole and brought with them two frontal passages and the first snow of the autumn.

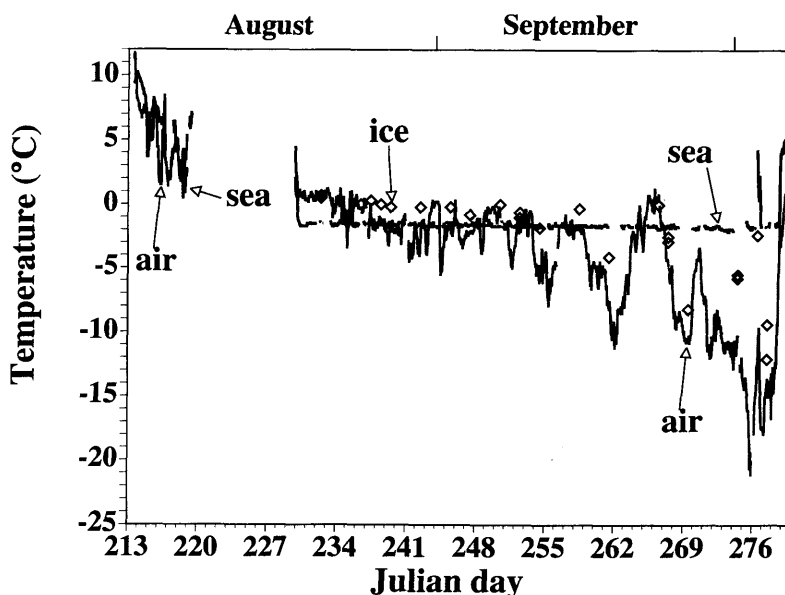


Fig. 2. Temperatures of air (30 m asl), surface water (3 m bsl) and ice surface (5 cm depth) measured from *Oden* during the IAOE-91.

Aged marine air masses, having a long residence time over the pack ice and subsidence became more frequent after mid-September. This in combination with the seasonal decrease in sunlight resulted in cooler air. The cloud cover and relative humidity decreased correspondingly but due to air temperatures below zero and occasional high relative humidities there were still mixing and radiation fogs of a more patchy and local character and sometimes thin stratus clouds. Visibility increased with the season, due to decreases in humidity and fog frequency. During the autumn the cloud cover consisted mainly of mid-level clouds (Altostratus and Altocumulus) and high clouds (Cirrus). Temperature and visibility decreased towards the pole while cloud cover and relative humidity increased.

The ice cover was limited to 60–95% during the summer part of the expedition by leads in the 1–4 m thick perennial pack ice but as the temperature decreased below the freezing point of sea water, especially from the middle of September, see Fig. 2, all leads were covered by new thin ice (Liljeström, 1992; Leck and Persson, 1996). The freezing of leads, snow fall and decrease in cloudiness resulted in a transition in albedo and the radiation and heat budgets. During autumn, from

having been a sink of heat, the sea became a source of heat and humidity through open leads as relatively cold air advected over the pack ice. For transition periods the ice itself was a source of heat, cf. Fig. 2. During clear sky conditions and negative surface radiation balance in late September and in October the closed ice isolated the air near the surface from the ocean source of heat, so that it became much colder than the air aloft. From 13 September (Julian day 256), when the sun began to set at night, the temperature decreased more rapidly and fell below -20°C . Inversions were caused by subsidence in anticyclones and occasionally at the surface by negative radiation balance in the absence of clouds.

During the autumn anticyclones become established but some cyclones entered the region and caused dramatic changes in weather. This caused large changes in temperature, humidity and cloudiness on the time scale of a few days, like the period of enhanced advection from the Atlantic Ocean in late September, see Fig. 2, during which the temperature increased to about 0°C . In total, only 5 fronts passed *Oden* during the expedition. In summary the weather pattern during the IAOE-91 was typical for the arctic summer and autumn.

3.2. Large-scale transport

The use of trajectories during the IAOE-91 is described in the appendix together with a validation of the trajectory model.

3.2.1. *Advection of marine air.* The time air masses spent over the pack ice or the time since they left the open sea (marine transport time, see the Appendix) increased with the season from summer to autumn, see Fig. 3 ($r = 0.255, p = 0.0015$, r is the correlation coefficient, p is the probability from regression analysis and an F -test that the correlation would occur by chance sampling fluctuations). This trend was interrupted by periods of increased transport from the Atlantic sector during late September. While *Oden* was in the pack ice, the median marine transport time was 52 h. This marine transport time correlated well throughout the expedition with a decline in air temperature ($r = -0.686, p \leq 0.0001$) and water vapour mixing ratio ($r = -0.662, p \leq 0.0001$) confirming that advection from the open sea is the central Arctic Ocean's main source of heat and moisture (Vowinkel and Orvig, 1970) in summer. This was of great importance for the formation of advection fogs and stratus clouds during the summer season, see Nilsson and Bigg (1996) in which also the transition from summer to autumn is described. The advection of heat and moisture will naturally be of great importance for the PBL.

3.2.2. *Subsidence and formation of continental air above the pack ice.* Trajectories revealed a change in vertical movements towards the end of the IAOE-91, roughly after 26 September (Julian day 269). Most subsidence took place south of the

ice edge at least 36 h ago for more than 90 % of the trajectories, but in the deep anticyclonic circulations that dominated the last ten days of the IAOE-91 subsidence took place north of the ice edge above the pack ice usually less than 48 h ago.

3.2.3. *Transport from surrounding continents.* The median crustal transport time (see the Appendix) was 91 h. Cases of transport from deep within the continents were rare and showed no significant trend. Events of transport from regions likely to be polluted were limited to only 6 % of the trajectories in total, concentrated in 2 episodes of transport from the Scandinavian peninsula, Finland, the Kola peninsula and the Arctic Russian coast. One of them is described in Bigg et al. (1996) and Covert et al. (1996). For the statistics of the air mass transport, see Table 1.

3.3. Structure and evolution of the arctic marine boundary layer

3.3.1. *The mixed layer.* A shallow Mixed Layer (ML) was found at the surface in about 85 % of the rawinsonde profiles typically 100–250 m in depth, as seen in Fig. 4b–d and Table 2. In it the virtual potential temperature θ_v had approximately uniform distributions. (Since $(\partial\theta_v/\partial z)$ was not always exactly zero in the ML we used a near neutrality criterion, $|\partial\theta_v/\partial z| < 5 \cdot 10^{-3} \text{ K m}^{-1}$, to define the height z_i of the ML). For air-sea exchange applications where the flux through the air-sea interface depends on the atmospheric stratification near the surface it should be reasonable to assume neutral conditions near the surface in the ML. The profiles with an ML present were related to trajectories that had travelled 400 to 4000 km over the ice. The profiles without an ML

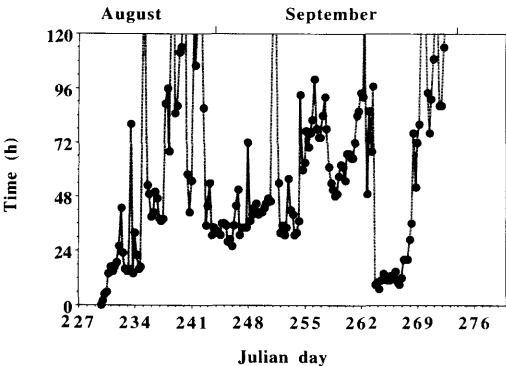


Fig. 3. The elapsed time of transport since open sea, marine transport time, during the IAOE-91.

Table 1. The descriptive statistics of transport times from marine, crustal and free troposphere sources during the IAOE-91

Source	Transport time (h)		
marine (in ice)	27	52	102
crustal	53	91	> 120
free troposphere	43	91	> 120
	1st quartile	2nd quartile	3rd quartile
		median	

The marine transport time is based exclusively on the period when *Oden* was in ice.

Table 2. *The characteristics of the planetary boundary layer during the IAOE-91*

	Min	Arithmetic mean	Geometric mean	Max	Range	σ	n
z_i (m)	15*	204	126	735	720	172	45
h' (m)	270	934	863	1700	1430	343	45
$\Delta\theta_v(z_i) - \Delta\theta_v(h')$ (K)	3	9.3	8.4	20	17	4.2	45
$H_{\Delta\theta}$ (m)	50	340	287	747	697	181	45

The mixed layer height z_i , the stable layer top h' , approximate strength $(\Delta\theta_v(h') - \Delta\theta_v(z_i))$ and scale height $H_{\Delta\theta}$. σ and n denote the standard deviation and number of cases, respectively.

* Mixed layers below the launch height of 15 m cannot be excluded.

either occurred during the first day of advection over the ice, at the most 325 km from the ice edge, cf. Fig. 4a, or had spent more than 5 days over the pack ice and occurred in conditions of clear sky and radiation inversions at the beginning of October.

Mixed layers over the central Arctic Ocean pack ice during spring, summer and fall have been previously described by Sverdrup (1930) as "cold layers" at the surface below an upper inversion, by Belmont (1956) and by Vowinkel and Orvig (1966) as an absence of surface inversions.

3.3.2. The stable layer and its structure. A Stable Layer (SL), with $\partial\theta_v/\partial z > 0$, was always present above the ML, cf. Fig. 4 and Table 2. The persistence of the SL is likely to be due to: negative radiation balance of the atmosphere, temperature advection and cooling at the surface of the relatively warmer air advected from the south, vertical differences in the speed of advection, negative radiation balance at the surface at the end of the expedition and subsidence in high pressure systems. An upper inversion was included in the lowest part of the SL in about 50% of the profiles, the inversion base separating the SL from the ML. When the SL started at the surface, which it did in 15% of the profiles, mainly near the ice edge in on-ice flow, cf. Fig. 4a, it can be regarded as a Stable Boundary Layer (SBL).

The SL height h is defined as the height at which $\partial\theta_v/\partial z$ (approximated with finite differences with $\Delta z = 10$ m) again equals 0. In most cases, $\partial\theta_v/\partial z$ slowly approached zero without ever reaching it. We therefore define h' as the height at which $\partial\theta_v/\partial z < 5 \cdot 10^{-3} \text{ K m}^{-1}$ became established. Uncertainties in h' are likely to be large and the larger vertical extent of the SL compared to the ML means that neither z_i nor h' offers a good

estimate of the bulk volume of the PBL. For these reasons a box model approach to the atmospheric chemistry and aerosol data of the IAOE-91 requires a scale height for the PBL. For this purpose we recommend the usage of an integral depth scale $H_{\Delta\theta}$ which we define as the approximated accumulated cooling of the SL, $\int_{z_i}^{h'} \Delta\theta_v(z) dz$, over the depth z_i to h' normalised by the approximated SL strength $(\Delta\theta_v(z_i) - \Delta\theta_v(h'))$:

$$H_{\Delta\theta} = \frac{\int_{z_i}^{h'} \Delta\theta_v(z) dz}{(\Delta\theta_v(z_i) - \Delta\theta_v(h'))}, \quad (1)$$

where $\Delta\theta_v(h') = \theta_v(h) - \theta_v(h')$, $\Delta\theta_v(z_i) = \theta_v(h) - \theta_v(z_i)$ and $\Delta\theta_v(z) = \theta_v(h) - \theta_v(z)$ in agreement with the notation used by Stull (1988) with the difference that we have the SL bottom at $z = z_i$ instead of 0. It was only rarely possible to estimate the total SL strength $\Delta\theta_v(z_i)$. In Fig. 5, examples of θ_v profiles are given with $(\Delta\theta_v(z_i) - \Delta\theta_v(h'))$, z_i and h' . The characteristic height suitable for box modeling is then $z_i + H_{\Delta\theta}$ and z_i and h' can be regarded as the worst possible error limits.

The SL thickness $h' - z_i$ correlated very well with the $H_{\Delta\theta}$ ($r = 0.925$, $p \leq 0.0001$). This linear relationship is consistent with both linear and exponential θ_v profile idealisations. However the high average ratio of $h' - z_i$ and $H_{\Delta\theta}$ of 2.3 supports a weak exponential θ_v profile idealisation

$$\Delta\theta_v(z) = \Delta\theta_v(z_i) e^{-(z - z_i)/H_{\Delta\theta}}, \quad (2)$$

with $H_{\Delta\theta}$ as the e -folding depth, while linear profiles always would have the ratio 2. Exponential schemes for idealised models of the long-term averages of SBLs have been used before by several authors (Stull, 1983; Carlson and Stull, 1986; Surridge and Swanepoel, 1987; Wong et al., 1987).

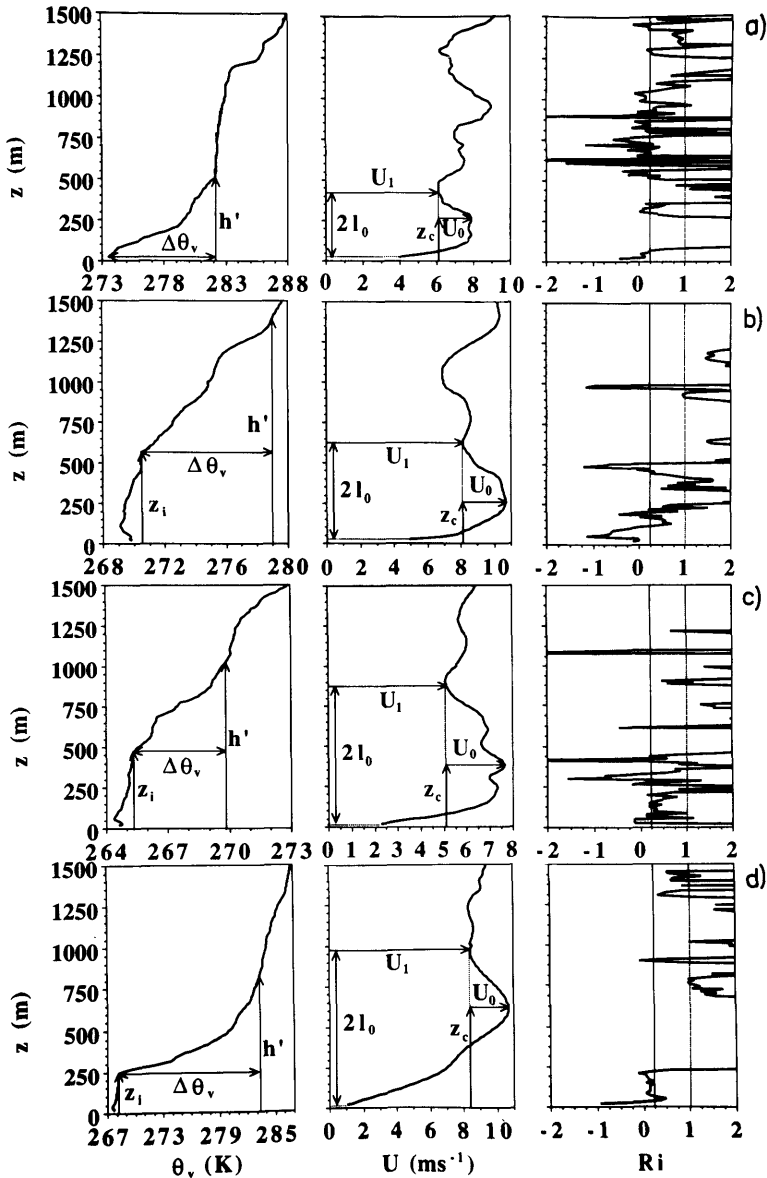


Fig. 4. Rawinsonde profiles of the IAOE-91. The mixed layer height z_i and stable layer scale height $H_{\Delta\theta_v}$, top h' , the low level jet centre z_c and scale height l_0 , free stream wind speed U_1 and scale of mean velocity variation U_0 are marked in the figures. (a) θ_v , U and Ri profiles at 16.11 UTC 20 August (Julian day 232) in air that had travelled about 275 km over the pack ice. Observe that there is no ML. (b) θ_v , U and Ri profiles at 03.10 UTC 12 September (Julian day 255) in air that had travelled about 1675 km over the pack ice. This launch is believed to have been made in a roll vortex. (c) θ_v , U and Ri profiles at 09.21 UTC 19 September (Julian day 262) in air that had travelled about 1500 km over the pack ice. (d) θ_v , U and Ri profiles at 18.07 UTC 24 September (Julian day 267) in air that had travelled about 475 km over the pack ice in an anti-cyclone resulting in a stronger SL with a subsidence inversion.

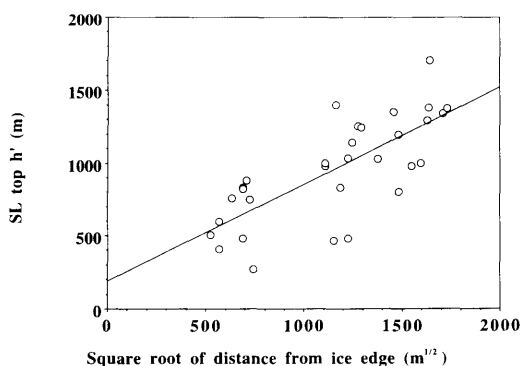


Fig. 5. The SL top h' against the square root of the fetch from the ice edge as revealed by rawinsondes and trajectories respectively during the IAOE-91.

Individual profiles may differ substantially from an exponential profile as seen in Figs. 4a–d. The average ratio between, the linear relationships between them and eq. (2) yield an approximated SL strength $(\Delta\theta_v(z_i) - \Delta\theta_v(h'))$ averaging 90% of the total SL strength $\Delta\theta_v(z_i)$. The unknown $\Delta\theta_v(z_i)$ in eqs. (1) and (2) can therefore be replaced with the known $(\Delta\theta_v(z_i) - \Delta\theta_v(h'))/0.9$.

The negative curvature (convex to the surface) of the SL θ_v profiles suggests that radiative and advective cooling dominated (Brunt, 1934; André and Mahrt, 1982). The weak to positive curvature in the ML confirms that turbulent heat transfer to the surface dominated there.

3.3.3. The vertical development above the pack ice. Intuitively, the height of the SL top should increase with time or distance north of the ice edge as the accumulated cooling of the air profile due to advection and heat flux into the ice and sea surface increases. To examine this, the 33 rawinsondes launched above the pack ice will now be considered. If two profiles during the last 10 days of the expedition are removed, the best correlation ($r = 0.742$, $p \leq 0.0001$) for h' with the square root of the distance x (fetch, (m)) from the ice edge is obtained:

$$h' \propto Cx^{1/2}, \quad (3)$$

where $C \approx 0.67$, as shown in Fig. 5. The 2 profiles that were excluded were obtained in air that had travelled the longest distance over ice, close to 4000 km, and experienced the strongest subsidence. The square root fits better with the data

than any other power function (including linear) or logarithmic function of either time or distance. The correlation is better for h' than for the SL thickness $h' - z_i$ since the accumulated cooling of the ML, $z_i \Delta\theta_v(z_i)$, must be included. The correlation is good considering the uncertainties in x , due to trajectory errors and the diffuse character of the ice edge (Leck and Persson, 1996), as well as the uncertainty in h' . Factors probably reducing the correlation are heat forcing from negative radiation balance in the SL in absence of clouds as well as profiles with subsidence inversions that have not successfully been excluded from the data set. If present, these processes started at other distances than at the ice edge. The IAOE-91 value for C in eq. (3) of 0.67 can be compared to values of 0.47, 0.57 and 0.51 obtained by Mulhearn (1981, interpreted by Hsu (1983)), Hsu (1983) and Hsu (1989) respectively.

This square root growth has previously been observed for SBLs or stable Internal Boundary Layers (IBL). Raynor et al. (1975) derived a function from experimental data, which included the square root of the fetch at off-shore flow over a sea surface colder than the air. Mulhearn (1981) derived a somewhat different expression which has later been given a physical explanation (Garratt, 1987; Garratt and Ryan, 1989):

$$h' = \alpha \bar{u} \left(\frac{g \Delta\theta}{\theta} \right)^{-1/2} x^{1/2}, \quad (4)$$

where $\bar{u}$ is the mean wind speed (in our case the mean wind speed along the trajectory), g the gravity acceleration, $\Delta\theta$ is the temperature difference between the surface and the upstream air or the air above the IBL top (in our case $\Delta\theta_v(z_i) - \Delta\theta_v(h')$) and α is dependent among other things on the θ_v and wind profile shapes.

Using eq. (4) and values of $\bar{u}$, $\Delta\theta_v(z_i) - \Delta\theta_v(h')$ and x from the IAOE-91 the correlation decreases ($r = 0.586$, $p = 0.0011$), although we again removed outliers with very large $\Delta\theta_v(z_i)$ due to subsidence. The slope of the best fit gives a value of $\alpha = 0.029$. This value is higher than the original value $\alpha = 0.014$ found by Mulhearn (1981) and derived by Garratt (1987) from modeled data. A higher value of α is in agreement with Garratt and Ryan (1989) ($\alpha \approx 0.024$, an average of values in the range 0.015 to 0.037) and Melas (1989) ($\alpha \approx 0.03$).

The IAOE-91 values of C and α in eqs. (3) and (4), respectively, agree fairly well with previous results, considering the larger uncertainty of x since the IAOE-91 observations are the first extending beyond 1000 km. Most of the previous work was valid for fetches <250 km, Garratt and Ryan (1989) extended to 560 km, and Hsu (1983) and Garratt (1987) to about 1000 km. It has been suggested (Mulhearn, 1981) that the profile should reach a self-preserving form at a sufficiently large distance for it to have reached equilibrium with the surface and ceased to grow. In our case, the growth rate after 3000 km would only be a fraction of the initial growth rate, so whether h' became constant or grew very slowly would make little difference. Also worth considering is that 3000 km corresponds to typically 2–3 days of advection. For eqs. (3) and (4) to apply for such a long period the boundary layer processes responsible for the growth, that all typically have shorter time scales, will have to last for the same period of time. It may be the more or less complete absence of diurnal variations in the arctic summer PBL that makes this possible.

3.4. The PBL structure of wind and turbulence

3.4.1. Low level jets. The average surface wind speed was 6.0 m s^{-1} (10.0 m s^{-1} at the SL top). Lower wind speeds were common and usually associated with high pressure circulations. The highest surface wind speeds exceeded 20 m s^{-1} and were related to strong transport from the Atlantic sector and cyclone activity. It was common for the wind speed to reach a maximum in the PBL, which will be referred to as a low level jet (LLJ).

We have normalised the LLJs to non-dimensional height η and non-dimensional wind speed $f(\eta)$ such that:

$$\eta = \frac{z - z_c}{l_0}, \quad (5a)$$

$$f(\eta) = \frac{U - U_1}{U_0}, \quad (5b)$$

where z_c is the LLJ centre, l_0 is the scale height of the LLJ, U_1 is the free stream wind speed at the LLJ top and U_0 the scale of mean velocity variation for the LLJ (Townsend, 1976). U_1 , U_0 , z_c and l_0 are marked in the examples of wind profiles of Fig. 4. LLJs were present in 90% of the profiles,

in 80% with $U_0 \geq 1 \text{ m s}^{-1}$ and in 60% with $U_0 \geq 2 \text{ m s}^{-1}$. More than one wind maximum was found in 30% of the profiles. The smoothing of the wind-profile, see Subsection 3.4.2, may have caused an underestimation of U_0 .

Fig. 6 contains the LLJs, normalised according to eqs. (5a, b), for which the scale height was found to be larger than the jet centre height ($l_0 > z_c$) due to the low vertical position of the LLJ near the top of the ML. They differed from the symmetrical LLJs found by Smedman et al. (1993) at higher levels in a stable MBL. The wind profiles below the level of maximum velocity had downward increasing shear and resembled hyperbolic cosine functions, while above it many profiles roughly followed a hyperbolic tangent function with an upward decreasing shear. Some were less regular but all had weaker shear above than below the velocity maximum. This difference resulted in an asymmetrical form. The initial uncertainty in the wind finding system (for typically the first two wind values) causes an uncertainty in the form of the lowest part of the wind profile. However, since the jet centre was usually well above this part of the profile and the wind at launch height was

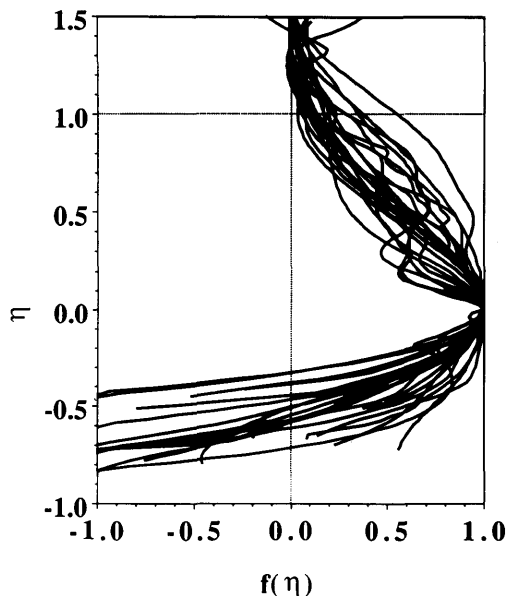


Fig. 6. Non-dimensional velocity $f(\eta)$ profiles plotted as a function of the non-dimensional height η , for low level jets with scale height larger than the jet centre height ($l_0 > z_c$) observed during the IAOE-91.

known, the conclusion that the average shear strength was stronger below the jet centre is unaffected. The characteristics of the IAOE-91 LLJs are summarised in Table 3.

Wind maxima in or near upper inversions in the high Arctic PBL were first described by Sverdrup (1930), who found them to be most common in winter and summer. More recently LLJs above the pack ice near the ice edge have been observed in winter and spring by Shapiro and Fedor (1986), Persson and Ruffieux (1992) and Brümmer et al. (1993). The observations during the IAOE-91 confirm the existence of LLJs far north of the ice edge in summer and autumn.

Langland et al. (1989) demonstrated with a two-dimensional model, initialised with winter data from the Spitsbergen ice edge in February (Shapiro and Fedor, 1986), that an ice-breeze mechanism is a possible cause of the LLJ near the ice edge. In on-ice flow an SBL formed at the ice edge with a depth that increased with the fetch and the LLJ appeared 100 km downwind from the ice edge. In off-ice flow the LLJ was formed at the top of an ML above the ice and was strongest near the ice edge. The similarities in temperature and wind profile structure between this study and that of Langland et al. (1989) are striking in spite of the difference in season and the limitation of the latter to not more than 200 km over the ice.

3.4.2. Layers of sub-critical Richardson number. To allow the use of the gradient Richardson number (Ri):

$$Ri = \frac{(g/\theta_v) \partial\theta_v/\partial z}{(\partial u/\partial z)^2 + (\partial v/\partial z)^2}, \quad (6)$$

where g is the gravity acceleration constant, z the height and u and v the wind components, and to minimise the effect of random experimental errors the profiles of the rawinsondes have been smoothed and cubic spline fitted in a least squares manner (NAG functions E02BEF and E02BBF). The derivatives of eq. (6) have been approximated with finite differences ($\Delta z = 10$ m). Ri relates the wind shear and the thermal stability. It is indirectly a relative measure of the rates of production/destruction of turbulent kinetic energy by buoyancy and wind shear. Theoretical work has shown that its critical value for the onset of turbulence in laminar flows is 0.25 (turbulence is continuous at values of $Ri < 0.25$). Empirical research supports this or a slightly lower value and suggests a critical value of approximately 1 for the termination of turbulence in turbulent flows that become laminar. At increasing Ri turbulence may continue at $Ri > 0.25$ until $Ri > 1$, while at decreasing Ri turbulence will not start until $Ri < 0.25$, so that there is a hysteresis effect for Ri between 0.25 and 1. The

Table 3. *The characteristics of the low-level jets found during the IAOE-91*

	Min	Arithmetic mean	Geometric mean	Max	Range	σ	n
all LLJs							
z_c (m)	63	447	345	989	926	302	55
l_0 (m)	44	312	264	924	880	183	55
U_0 (m s ⁻¹)	0.5*	2.6	1.8	10.1	9.8	2.3	55
U_1 (m s ⁻¹)	0.8	8.0	6.7	26.2	25.4	4.7	55
z_c/z_i			2.4				
LLJs with $l_0 > z_c$							
z_c (m)	63	209	187	430	367	101	25
l_0 (m)	101	276	250	490	389	123	25
U_0 (m s ⁻¹)	0.8	3.0	2.5	10.1	9.3	2.3	25
U_1 (m s ⁻¹)	0.8	7.0	5.4	15.8	15	4.2	25
z_c/z_i			1.6				

The low-level jet centre z_c and scale height l_0 , scale of mean velocity variation U_0 and free stream wind speed U_1 , σ and n denote the standard deviation and number of cases respectively.

* The existence of wind maxima weaker than σ_U cannot be excluded.

theoretical and empirical work in the field have been summarised by Arya (1988) and Stull (1988).

Besides the ML, where subcritical values of Ri and hence turbulence were common, layers of subcritical Ri were frequently found within the shear of the LLJs as seen in Figs. 4a–d. They occurred in 80% of the profiles with $Ri < 1$ and in 65% with $Ri < 0.25$. In Fig. 7, the frequency of subcritical Ri is plotted against η . Maxima in frequency of subcritical Ri , and hence turbulence, were found in two regions around $\eta = 0.2$ and $\eta = -0.6$, above and below the LLJ centres respectively, with a minimum in between at $\eta = 0$, which agrees with the structure observed by Marth et al. (1979) in nocturnal SBLs. At the LLJ top and bottom, $\eta = \pm 1$, the frequency of subcritical Ri approached the low values generally found in the SL, which suggests that much of the turbulence within the SL was directly associated with the shear of the LLJs. The fact that the frequency was higher below than above the LLJ centres is due to the asymmetrical LLJs in Fig. 6, with their stronger wind shear below the LLJ centres in the more or less continuously turbulent ML.

3.4.3. Evidence and possible explanations of intermittent mixing. Large changes in Arctic PBL temperature profiles in short intervals of time have been observed (Evgenow, 1931; Cheung, 1991). Cheung (1991) observed a sudden change over half an hour in air temperature profiles in the Arctic lower PBL and sodar observations were able to relate this to entrainment of warmer air from

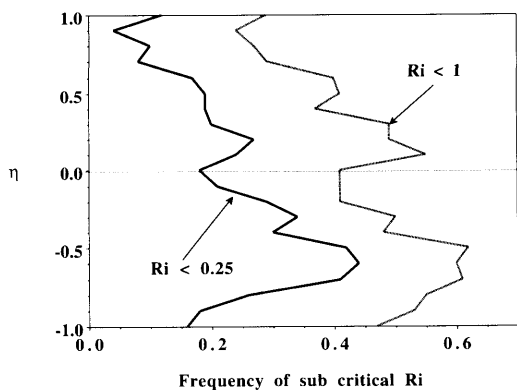


Fig. 7. The frequency of subcritical Ri ($Ri < 0.25$ and $Ri < 1$), and hence turbulence, as a function of the non-dimensional height η for all low level jets observed in wind profiles of rawinsondes during the IAOE-91.

above, aided by breaking waves. Changes on the same short time scale were frequently observed during the IAOE-91 in air temperature as well as in aerosol and tracer concentrations (Bigg et al., 1996). Nilsson and Bigg (1996) suggested that perturbations caused by shear induced waves sometimes influenced fog formation and dissipation during the IAOE-91. The stability of the SL will preserve gradients in tracer, aerosol or water vapour concentrations. If mixing took place intermittently in the SL and reached the ML it could explain the observed sudden changes in air temperature and tracer concentrations by mixing of two or more layers of air that due to their stability may not only have different temperatures but also different concentrations of tracers. There are several possible reasons for the intermittent mixing. The most likely candidates are probably Kelvin-Helmholtz (K-H) waves, shear induced gravity waves and the intermittent PBL breakdown phenomena. These will now be briefly discussed and support for their presence during the IAOE-91 will be examined.

The wind shear of the LLJs in the layers of subcritical Ri identified above ($Ri < 0.25$) can result in K-H waves that grow in amplitude and eventually begin to roll up and break. As a wave breaks its energy is lost to turbulence reducing shear and stability. See Stull (1988) for a summary. If this happens just above the ML it would mix down air from the lowest part of the SL to the ML. K-H waves were observed as billow clouds, at the top of fog or thin and low stratus, since droplets acted as visible tracers and condensed and evaporated in the rising and sinking portions of the wave respectively. This, together with the high frequency of observations of layers of subcritical Ri , suggests that K-H waves were indeed present in the PBL during the expedition and that they may have been a common phenomenon.

Layers of subcritical Ri like those in Fig. 4 are not only favourable for K-H waves but also for shear induced gravity waves. The shear of the LLJ, or according to Stull (1988) the K-H waves, may cause internal gravity waves. These may propagate vertically from the shear layer if sandwiched between stably stratified regions. If a gravity wave radiates downwards mixing between the SL and the ML may result.

In spite of the smooth terrain and slowly changing environment of the central Arctic Ocean PBL,

wave phenomena have been observed there (Cheung, 1991; Walter and Overland, 1991; Brümmer et al., 1994).

Intermittent PBL breakdown or collapse can be caused by transitions between laminar and turbulent flow, non-critical and subcritical Ri , in a near equilibrium between radiative and turbulent heat flux (Blackadar, 1979; ReVelle 1993). According to ReVelle (1993) the process is favoured by a narrow range of geostrophic winds and surface roughness lengths. This is a possible explanation of the observed periodicities, or at least individual sudden changes, although the modeled intervals of 2–4 h is longer than the periodicities observed during the IAOE-91 in aerosol and radon concentrations of typically 1.0–1.5 h (Bigg et al., 1996).

3.5. Roll vortices above pack-ice

Secondary circulations during cold air outbreaks from the pack ice over adjacent open seas have received quite a lot of attention (Walter, 1980; Hein and Brown, 1988; Chlond 1992; Brümmer et al., 1992). Although cloud streets above the Arctic pack ice had been observed as early as 1957 (Kuettner, 1959), little work has been done on secondary circulations above ice (Walter and Overland, 1984) and above the permanent pack ice only in the ice edge zone (Brümmer et al., 1994). It was not until after the expedition that the possibility was realised that they could strongly influence cloudiness and aerosol concentrations. Consequently the evidence presented below is derived from information available after the expedition rather than planned observations.

3.5.1. Cloud streets. On the NOAA 11 satellite images linear cloud patterns approximately parallel to the wind direction were found frequently. In the example of Fig. 8, cloud streets can be seen not only over open sea but also well into the pack ice. This picture has been cut to show mainly the area of clearly visible cloud streets. At the position of *Oden*, just north of the picture, high clouds obscured the lower clouds. The average roll wave length λ_R of the cloud streets was estimated in a way similar to Walter (1980). Using all satellite pictures available from the expedition this resulted in λ_R of 4–12 km. The cloud streets covered areas up to tens of thousands of km^2 . They were often visible in channels 1 and 2 (visible light) but not in channel 3 (infrared). The latter channel is the one that most clearly shows all low clouds

and fogs however thin, while the shorter wavelengths penetrate the thinnest clouds. Hence we arrive at the conclusion that the roll vortices frequently caused alternating bands of thinner and thicker clouds rather than distinct cloud streets.

3.5.2. Properties of the roll vortices. Fig. 9 shows the wind rotated into mean ML wind coordinates of a rawinsonde launched on 12 September (Julian day 255), cf. Fig. 4b. This profile has been discussed in an accompanying paper (Bigg et al., 1996) in connection with periodicities in the aerosol number concentration. It comes from the same day as the satellite picture of Fig. 8, although high clouds obscured the lower clouds at the position of *Oden*. The mean ML wind direction (lateral) should be close to the direction of the roll axis. The profiles provided the following information: $z_i = 560$ m, the inversion top $z_t = 700$ m, an inflection point of the lateral wind was present at $z_{ip} = 300$ m, mean ML wind direction was 166° ; the relative humidity was $\geq 90\%$ from 315–680 m, the stratus cloud top; and the wind direction was 186° at the SL top. The lateral wind profile resembles quite well a modified lateral Ekman wind profile containing roll vortices (Brown, 1970; Hein and Brown, 1988). The angle between the roll axis direction and the geostrophic wind α_R estimated to be 20° agreed well with Brown's (1972) prediction for a near neutral stratification.

The rawinsondes of the IAOE-91 suggested values within the entire predicted range of α_R . The wind profile above the SL, the relatively colder air temperature of -3.2°C at 15 m asl compared to the sea and ice surface temperatures of -1.7°C and -1.4°C , respectively, and the northwest origin of the trajectories agrees with cold air advection over a warmer surface.

On 24 September (Julian day 267) at 11:55 UTC, a satellite picture was received with cloud streets above and around *Oden* with $\lambda_R \approx 5.4$ km. Unfortunately the picture was not of printable quality. A rawinsonde launched at 10:10 UTC gave an estimated value of the roll height h_R of 730 m, which resulted in an aspect ratio λ_R/h_R of 7.4. For the IAOE-91, h_R could only be directly related to a value of λ_R for this single case, since no other satellite photos with cloud streets visible above *Oden* happened to coincide with a rawinsonde launch. However, it is consistent with the few earlier observations above ice of an aspect ratio between approximately 7 and 9 (Walter,

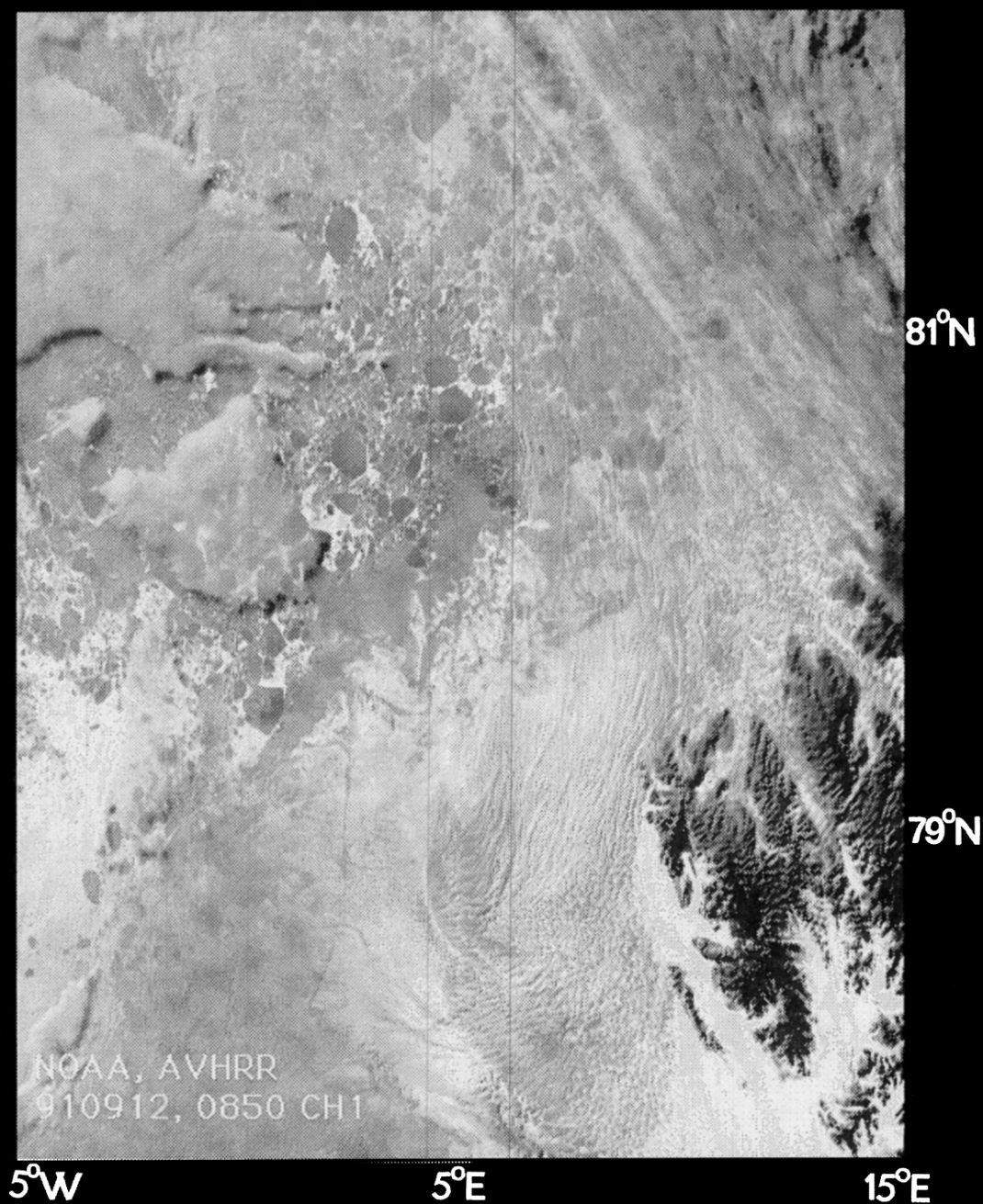


Fig. 8. NOAA-11 satellite image, channel 1 (visible light) over the northwestern corner of Spitsbergen, the Fram Strait and parts of the central Arctic Ocean, 08.50 UTC 12 September 1991 (Julian day 255) during the IAOE-91. Cloud streets can be seen both over open sea, in the marginal ice zone and well into the central Arctic Ocean pack ice.

1980; Walter and Overland, 1984; Hein and Brown, 1988) if we use the shortest wavelength observed on satellite images. Above aspect ratios of 7 mesoscale convective cells rather than roll vortices should be formed (Mitchell and Agee, 1977). High aspect ratios may be explained by irregularities in the roll structure or if the SL above forces the rolls to become flattened (Walter and Overland, 1984).

3.5.3. The organized mixing in roll vortices. Since the roll axes are tilted relative to the geostrophic wind direction, the up- and downdraft regions will pass any fixed point at the surface at intervals. Smedman (1991) has observed enhanced turbulence in updraft regions with such intervals of 10 to 15 min for relatively small roll vortices in a shallow PBL. The observed ranges of λ_R , α_R and the geostrophic wind U_g (approximated with the wind at the PBL top) during the IAOE-91 were such that the expected interval, estimated as the ratio between λ_R and the lateral wind component based on α_R and U_g ($\approx \lambda_R / (U_g \sin \alpha_R)$), could be anything from 0.2 h to more than a day. When U_g and α_R were averaged, the preferred period becomes 0.5–1.4 h. This range of periods agrees fairly well with the favoured periodicities of about 1.0–1.5 h (the rawinsonde of Fig. 9 was launched during a 1.6 h periodicity) in radon and aerosol number concentrations observed during the IAOE-91 (Bigg et al., 1996).

The presence of roll vortices may be of great importance for cloud and aerosol physics and chemistry, not only because they enhance mixing, increase z_i , entrain air from above the cloud top and influence cloud formation, but also because they drive air through many cycles of condensation and evaporation. With velocities of typically up to 1 m s^{-1} the "turnover time" of one roll cycle would have been a few hours during the IAOE-91. Nilsson and Bigg (1996) have related roll vortices to the periodic formation of fog like the fog streets observed by Walter and Overland (1984).

3.5.4. Indications of the cause of roll vortices. There are 2 basic types of instability that can lead to secondary flows in the PBL: dynamic and convective, and a review of them has been given by Brown (1980). The presence of an inflection point in the lateral wind component is a necessary condition for instabilities of the mean flow and a ratio between the inflection point height z_{ip} and the inversion top z_t around 0.5 is favourable

(Brummer and Latif, 1985). A small disturbance will then grow if the lateral wind shear, and hence the Reynolds number Re , is large enough (Lilly, 1966). Convective instabilities could be caused by heating at the surface as air aged over the pack ice is advected from cold surfaces with frozen ice floes and leads towards areas with a warmer surface, wet ice floes and open leads. Such conditions were especially common during the freeze-up of the central Arctic Ocean in the second half of September, cf. Fig. 2. Other processes that could cause convection in Arctic roll vortices is cloud top cooling and release of latent heat from condensation (Chlond, 1992).

We have already seen evidence of both dynamic and convective instabilities in the example of Fig. 9. Assuming that the periodicities in radon and aerosol concentration (Bigg et al., 1996) were caused by roll vortices, we compare rawinsonde profiles from such with the other profiles. All profiles of the first group had an inflection point and z_{ip}/z_t near 0.5. As a group they had higher z_i and they were significantly (95% or better) different, according to a t -test, from the other group considering the air to surface temperature difference and the fetch. The first group had colder air with averages of the air temperature minus sea and ice surface temperatures of -4.3 and -2.5°C , respectively, and it had longer fetch distances, with an average of 2100 km, compared to the other group. This indicates that the roll vortices during

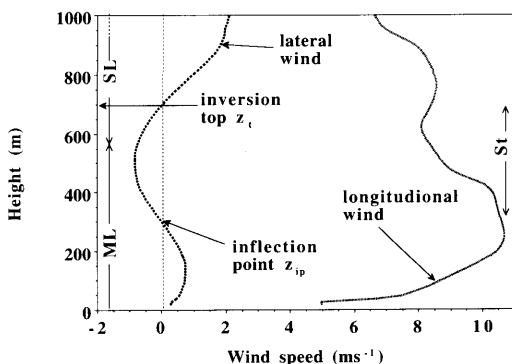


Fig. 9. The lateral and longitudinal wind speeds, relative to the mixed layer mean wind, of a rawinsonde launched from Oden 03.10 UTC 12 September (Julian day 255) during the IAOE-91. The mixed layer, stable layer, stratus (St) layer, inversion top z_t and the inflection point z_{ip} are indicated.

the IAOE-91 were driven by a combination of dynamic and convective energy, and that they appeared in air aged over the ice. Favourable lateral wind profiles, cloud streets and periodicities in radon and aerosol concentrations were indeed most common during the second half of September.

4. Summary and conclusions

The present paper, having established that weather and large-scale transport during the expedition were typical for the time of the year, has concentrated on the structure and evolution of the PBL and the disturbances that possibly affected it.

Because the expedition spent an extended time north of the ice edge it was possible to trace the evolution of the PBL, schematically described in Fig. 10, due to heat losses at the surface as air was advected over the ice for distances up to about 3000 km. The increase in depth of the SL proportional to the square root of the distance from the ice edge was known from earlier empirical and theoretical work, but has not been verified experimentally up to more than 1000 km, and the formation of an ML at the surface after 400 km has not been reported before. The suggestion that the wind shear of LLJs is involved in the formation and maintenance of the ML appears very plausible but certainly requires further study. The evidence also seems to point strongly to the importance of the LLJ and associated layers of subcritical Ri number in generating intermittent mixing events. As the

cooling at the surface ceased and the heat flux changed sign, the increase of the SL top ceased and weak convection became possible. Roll vortices, preferentially formed in air aged over the pack ice, caused organised mixing and increased the ML depth, but the overall PBL structure remained.

The distance from the ice edge at which the heat flux changes sign will depend on initial conditions of the air, surface conditions and on the season. In the summer months June through August when advection brings warm marine air it may not always happen. For periods during spring and autumn, when the ice can be warmed by the sun in the absence of complete cloud cover and especially during the autumn when heat is stored for a while in the ice, no significant downward surface heat flux at all has to occur and roll vortices may start earlier.

This evolution of an initially stable MBL needs to be further studied to decide after what distance and why the ML forms and at what distance and under what conditions the roll vortex circulations start. Also the mechanisms of the LLJ, the associated intermittent mixing and the mechanisms behind the roll vortices above ice call for further attention.

One of the main aims of the atmospheric programme during the IAOE-91 was to study the fluctuations in concentrations of sulfur gases in the Arctic MBL and to assess their contribution to the local aerosol. Overlying clouds are certain to have influenced reactions of the gases and to have acted as both sources and sinks of aerosols. Vertical mixing therefore allows the clouds to influence

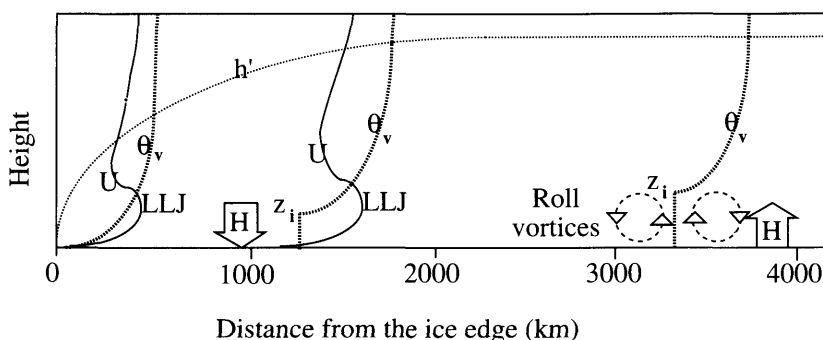


Fig. 10. The evolution of the marine boundary layer with increased fetch as air is advected into the central Arctic Ocean pack ice. In the figure z_i indicates the mixed layer height, h' and the thin dotted line marks the stable layer top, full line and thick dotted line shows idealised profiles of U and θ_v respectively. LLJ marks the low level jets. Arrows with H indicates the direction of the surface flux of sensible and latent heat.

the entire PBL. All chemical and aerosol physical measurements were necessarily made near the surface and would therefore have been very strongly influenced by intermittent or organised mixing that brought air from higher levels.

The evidence for the mixing phenomena relies heavily on stability and wind profiles in conjunction with occasional satellite images in the case of roll vortices. A more direct method of detection such as ultrasonic anemometers or acoustic sounders would be of immense benefit in verifying the nature of the disturbances involved when sudden changes in temperature, wind speed, relative humidity, aerosol and trace gas concentrations occur. If vertical profiles of aerosol and gas concentrations could be obtained, fluctuations in surface concentrations would provide details of how mixing occurred.

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6. Appendix

6.1. The use of trajectories during the IAOE-91

For the period when *Oden* was in the pack ice, the calculated trajectories were compared with ice

maps and estimates made of the elapsed time of transport from the open sea (the marine transport time) as well as the distance travelled by the trajectories (the fetch x). Trajectories and maps gave the transport time since the last contact with snow- and ice-free land and crustal sources (crustal transport time), continental or any of the islands, not ice covered, in and surrounding the central Arctic Ocean. Furthermore we calculated the elapsed time since the trajectories left the free troposphere and entered the PBL using trajectories, weather maps and rawinsoundings. This approach is very useful because it can be used to sort other data efficiently and objectively according to the air mass history. It can also be used together with meteorological, chemical or physical data to study how parameters evolve with time in a Lagrangian manner.

6.2. The validity of the trajectory model during the IAOE-91

Due to the limited number of observations of the mass and wind fields in the high Arctic it is necessary to examine the accuracy of the trajectory model. This has been done by combining the transport time from the three categories of sources, marine, free troposphere and crustal, with gas chemistry tracers, dimethylsulfide (DMS), ozone (O_3) and radon (Rn) originating from these sources respectively. The open sea is not only a source of heat, moisture and sea-salt aerosol but also of the biogenic gas DMS. In remote regions O_3 has a net source to the PBL from the free troposphere. Rn is a radioactive decay product of radium and originates from the continents.

In the high Arctic Rn to a large extent originates from islands and coasts within or surrounding the Polar Ocean. These are almost point sources. The radon half-life of 3.8 days however means that strong distant sources may also exert an appreciable influence. In contrast, DMS with its ocean origin has large area sources or at least wide line sources along the ice edge within the Arctic ocean (Leck and Persson, 1996). To estimate how well the trajectory model resolves sources of different type and areal size and at different distances the correlations (r) for transport time of increasing time spans with the logarithm of the tracer concentration have been calculated, as shown in Table 4. The logarithm of the concentration is used since this corresponds to first order sinks, which is

Table 4. Correlations between elapsed time of transport from different sources and the logarithm of tracers concentration typical for these sources and how the correlation developed with increasing time interval of the trajectory model's backward calculation

Source	Marine				Crustal		Free troposphere			
tracer	DMS				radon		ozone			
time span	instantaneous		instantaneous		averaged ± 3 h		instantaneous		averaged ± 3 h	
(h)	<i>r</i>	<i>p</i>	<i>r</i>	<i>p</i>	<i>r</i>	<i>p</i>	<i>r</i>	<i>p</i>	<i>r</i>	<i>p</i>
0–24	–0.32	0.065		*		*		*		*
0–36	–0.33	0.033		*	–0.44	0.068		*		*
0–48	–0.40	0.004	–0.39	0.018	–0.50	0.001		*		*
0–60	–0.37	0.005	–0.29	0.036	–0.39	0.004		*		*
0–72	–0.44	0.000	–0.23	0.060	–0.26	0.030		*		*
0–84	–0.48	0.000	–0.20	0.064	–0.19	0.073		*	–0.22	0.028
0–96	–0.56	0.000	–0.21	0.034	–0.18	0.062		*	–0.26	0.005
0–108	–0.54	0.000	–0.16	0.083		*	–0.29	0.003	–0.37	0.000
0–120	–0.55	0.000	–0.15	0.091		*	–0.38	0.000	–0.45	0.000

r and *p* denote the correlation coefficient and the probability that the correlation would occur by chance from an *F*-test.

* Probability $p > 0.1$.

accurate at least for DMS and Rn. Crustal sources had their maximum correlation with Rn in the interval 0 to 48 h transport time. Beyond 48 h the correlation decreased and the probability (*p*), from regression analysis and a *F*-test, that the correlation would occur by chance sampling fluctuations approached 0.1 after more than 96 h of transport time. The marine sources on the other hand correlated well throughout 120 h. This implies that the trajectory model could resolve large surface sources five days or more backward in time, while the resolution of limited area or point sources becomes uncertain beyond four days.

For ozone of free tropospheric origin the correlation was better ($r = -0.451$, $p < 0.0001$) if

the concentrations were averaged, ± 3 h from the arrival time of the trajectories, than for the instantaneous concentrations, as shown in Table 4. Radon from the crustal sources showed the same behaviour during the first 72 h and most of the improvement appeared already if averaging over ± 1 h. The fact that the correlation for averaged concentrations of ozone and radon was better than the instantaneous indicates that there were variations in concentrations on a time scale less than 6 h that the trajectories were not able to resolve. Such variations were indeed observed in radon, ozone and aerosol number concentrations (Bigg et al., 1996) during the IAOE-91 on a time scale of about one hour and they were related to boundary layer processes discussed in this paper.

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